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Research paper

Fault on-off versus strain rate and earthquakes energy

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A B S T R A C T

We propose that the brittle-ductile transition (BDT) controls the seismic cycle. In particular, the movements detected by space geodesy record the steady state deformation in the ductile lower crust, whereas the stick-slip behavior of the brittle upper crust is constrained by its larger friction. GPS data allow analyzing the strain rate along active plate boundaries. In all tectonic settings, we propose that earthquakes primarily occur along active fault segments characterized by relative minima of strain rate, segments which are locked or slowly creeping. We discuss regional examples where large earthquakes happened in areas of relative low strain rate. Regardless the tectonic style, the interseismic stress and strain pattern inverts during the coseismic stage. Where a dilated band formed during the interseismic stage, this will be shortened at the coseismic stage, and vice-versa what was previously shortened, it will be dilated. The interseismic energy accumulation and the coseismic expenditure rather depend on the tectonic setting (extensional, contractional, or strike-slip). The gravitational potential energy dominates along normal faults, whereas the elastic energy prevails for thrust earthquakes and performs work against the gravity force. The energy budget in strike-slip tectonic setting is also primarily due elastic energy. Therefore, precursors may be different as a function of the tectonic setting. In this model, with a given displacement, the magnitude of an earthquake results from the coseismic slip of the deformed volume above the BDT rather than only on the fault length, and it also depends on the fault kinematics.

1. Introduction

An earthquake occurs when the volume close to a fault moves, the "fault on" stage. This coseismic stage is preceded by a long-lasting interseismic stage, called here as the "fault off" period. This evolution is commonly known as the seismic cycle (e.g., Thatcher and Rundle, 1979; Savage, 1983; Cattin and Avouac, 2000; Meade and Hager, 2005; Sieh et al., 2008). Faults are locked or unlocked depending on a number of parameters, including the presence of asperities and elasto-plastic instabilities (Kanamori and Anderson, 1975; Ruina, 1983; Hobbs and Ord, 1988; Marone, 1998).

In this paper, we consider the partitioning of strain between brittle and ductile structural levels as one of the mechanisms controlling the seismic cycle. The terms brittle and ductile could be substituted with Sibson's (1977) classification, where the elastic-frictional deformation dominates within the upper crust, and quasi-plastic deformation prevails in the lower, warmer crust. The terms brittle and ductile are used in the following for the sake of simplicity. The brittle behavior corresponds to frictional failure, whereas ductile deformation corresponds to thermally-activated creep. Faults and the related deeper shear zones evolve by different processes at different depths. A fault mylonite at depth corresponds to a cataclasite at shallower depth. At the transition between these two deformation mechanisms, viscous processes leave space to elastic-frictional processes at shallower depth. Although Sibson (1977) more precisely refer to the rheological processes, here we give emphasis to the simplicity of the fault activation model and we rely on a two-layer crust with a single brittle-ductile transition (BDT), mainly derived from Doglioni et al. (2011). Our model links the ongoing viscous deformation at depth with the fragile-brittle episodic behavior in the upper crust. We are clearly aware that this is an over-simplification since more than one

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BDT may occur in the crust as a function of the embedded lithologies. The results illustrate how the BDT focuses the energy loading in the overlying volume of rock during the interseismic stage. With the proposed rock deformation mechanics, the BDT and its depth control the energy accumulation, strain-rate distribution, and fault movement. In particular, we suggest that the strain rate gradients indicate stress accumulation. We compare the results of numerical models (Doglioni et al., 2011, 2013) with GPS data in a number of seismic crises in northern America, Chile, Italy, Taiwan and Turkey. Finally, a discussion on the different energy storage during the interseismic period is discussed as a function of the tectonic style.

2. Geological model of fault activation

The BDT represents the lower limit for most of the crustal seismicity worldwide (Scholz, 1990, and references therein), and it appears to have a relevant role in the description of the seismic cycle (Doglioni et al., 2011). The BDT separates the seismogenic elastic-frictional upper layer from the underlying quasi-plastic layer. In this scenario, stress in brittle upper crust builds up during the interseismic period until it ruptures seismically and deforms instantaneously reversing their relative motion along the conjugate bands formed above the BDT (Doglioni et al., 2011). We hypothesize an active fault crossing the entire crust exhibiting episodic locking-unlocking behavior in the brittle upper crust and a constant strain rate along the shear zone in the ductile lower crust (Fig. 1). At the BDT, the strain transfers from one zone to the other. During the interseismic period, the absence of dislocation in the brittle layer pairs with the continuous slip in the ductile layer and generates a stress gradient at the BDT. The stress gradient eventually dissipates, dislocating the brittle segment during the coseismic stage. In our model, the contrasting deformation style across the BDT acts as an “off” (locked or no displacement) and “on” (unlocked or active displacement) switch controlling the seismogenic behavior of the fault in the brittle layer.

2.1. Normal fault

In normal faulting, deep ductile deformation generates a dilation band nucleating from the BDT interface within the hangingwall volume, which is antithetic to the brittle, shallow, locked portion of the fault (Fig. 1). Although the lithostatic load increases with depth, the dilated band is inferred to expand at depth while approaching the BDT, and it should form for the differential motion between the steadily shearing lower ductile layer with respect to the locked upper brittle part of the hangingwall (Doglioni et al., 2011).

Dilatancy occurs in this tensional band, i.e., fractures and cracks form and open when rocks are stressed (e.g., Frank, 1965). The width of the dilated area (where fractures and cracks form) ranges from tens to hundreds of meters, up to the kilometric scale (Doglioni et al., 2011), depending on the mechanical properties of the rocks, which, in turn, vary as a function of the lithology and physical conditions (T, P, fluid pressure). During the interseismic period, the dilated volume may gradually expand to generate a band progressively weakened by fracturing, which allows (or depends on) fluid migration. Interseismic and preseismic fractures are expected to form parallel to the fault plane (and to σ_2), and with both antithetic and synthetic dip in the hangingwall volume above the BDT, particularly along the antithetic band. This weak band is predicted to form in the brittle layer where at the surface a strain rate gradient is observed. At the coseismic stage, the weak band could be activated as an antithetic normal fault (Fig. 1). For the L’Aquila 2009 event, the antithetic band activated as a conjugate normal fault may be located at the western border of the subsided area (as mapped by Atzori et al., 2009, by means of InSAR data), where the strain rate was lower during the interseismic stage (Fig. 2). A second example is represented by the antithetic normal fault that slipped during the 40-s sub-event of the Irpinia 1980 earthquake in southern Italy (Pingue and De Natale, 1993). During the coseismic stage, the weight of the brittle hangingwall overcomes the strength of the weakened dilated band (Fig. 1). The triangle suddenly falls when the shear stress along the locked part of the fault overtakes the strength of the stretched band above the BDT. The hangingwall of the main fault instantaneously subsides closing the fractures in the dilated volume (Doglioni et al., 2011). The distribution of foreshocks and aftershocks of the L’Aquila earthquake agrees with this model in that foreshocks concentrated primarily along the dilated area above the BDT. The aftershocks (Chiarabba et al., 2009; Di Luccio et al., 2010) were rather well aligned along the activated upper segment of the normal fault and within the fallen hangingwall.

A field example of an outcropping paleo-BDT is the Permo-Liassic Lugano normal fault in the southern Alps, Italy (Bernoulli, 1964). The Alpine shortening exhumed such a fault, which can be analyzed in its different ductile and brittle portions. Metamorphism and shear is documented in the ductile deeper section at a paleodepth of 8–10 km (Bertotti et al., 1993) whereas several cracks occur in the brittle shallower part of the listric fault. Nuchter and Stockhert (2008) described cracks and veins, located close to the BDT. They proposed their generation during the coseismic stage. We wonder whether this relevant observation could fit the generation of veins also during the interseismic period, given the constraint that the overburden does not change significantly during vein formation. If this were the case, the veins would represent an exhumed analog of the dilated band forming during the interseismic stage at depth above the BDT, conjugate to the normal fault.

2.2. Thrust fault

Unlike the normal fault case, the hangingwall above the BDT of a thrust fault is compressed during the interseismic stage, accumulating elastic energy, and dilates at the coseismic stage, dissipating the energy stored (Fig. 1). In the Tohoku-Oki 2011 earthquake, the stress change suggests an active slip of several tens of meters along the plate interface during the coseismic fault weakening, a nearly total stress drop (Lin et al., 2013). The capacity to accumulate this energy depends on the rheological properties of the hangingwall. During the coseismic stage, elastic rebound is expected to generate uplift of the hangingwall above the brittle segment of the thrust fault and internal subsidence above the BDT, where some dilatancy is predicted to occur instantaneously in the area/volume previously overcompressed during the interseismic stage (Fig. 1). This model is consistent with the data and model presented by Burrato et al. (2003) for fault-propagation folding. A similar behavior was detected during the great 2004 Sumatra earthquake (Meltzner et al., 2006).

At the scale of an orogen, a similar evolution can be predicted. Along subduction zones, the slab hinge can either migrate toward or away from a reference point pinned on the upper plate (Doglioni et al., 2007). Along the Taiwan, Andean, Sumatra and Cascadia subduction zones, the hinge migrates towards the upper plate (i.e. eastward or northeastward), and the associated orogens undergo compression while the plate is moving westward or southwestward relative to the lower plate (Fig. 3). In these subduction zones, the convergence rate is accommodated partially by the subduction, and partially by upper plate shortening (Fig. 3). For example, the hinge of the Andean subduction zone is converging toward the South American upper plate (e.g., Doglioni et al., 2007). During the interseismic period, the steady movement of the GPS sites can be inferred as the superficial record of ductile deformation in the lower crust. Shortening is accommodated by stationary or steady